Invited review

Deep crustal expressions of exhumed strike-slip fault systems: Shear zone initiation on rheological boundaries

Shuyun Cao a,b,⁎, Franz Neubauer b

a State Key Laboratory of Geological Processes and Mineral Resources, Center for Global Tectonics and School of Earth Sciences, China University of Geosciences, Wuhan 430074, China
b Dept. Geography and Geology, University of Salzburg, Hellbrunnerstr. 34, A-5020 Salzburg, Austria

A B S T R A C T

The formation of major exhumed strike-slip faults represents one of the most important dynamic processes affecting the evolution of the Earth’s lithosphere. Detailed models of the potential initiation, their properties and architecture of orogen-scale exhumed strike-slip faults, which are often subparallel to mountain ranges, are rare. The initiation of strike-slip faults is at depth, where temperature-controlled rheological weakening mechanisms play the essential role localizing future strike-slip faults. In this review study, we highlight that in pluton-and metamorphic core complex (MCC)-controlled tectonic settings, as end-members, the initiation of strike-slip faults occurs by rheological weakening along hot-to-cold contacts deep within the crust and mantle lithosphere, respectively. These endmember processes are potential mechanisms for the initiation of orogen-scale exhumed strike-slip faults at depth resulting in a specific thermal and structural architecture. Similar processes guide the overall displacement and ultimately the exhumation at such deep levels. These types of exhumed strike-slip dominated fault zones expose a wide variety of mylonitic, cataclastic and non-cohesive fault rocks on the surface, which were formed at different structural levels of the crust during various stages of faulting and exhumation. Exhumation of mylonitic rocks is, therefore, a common feature of such reverse oblique-slip strike-slip faults, implying major transtensive and/or transpressive processes accompany pure strike-slip motion during exhumation. A major aspect of many exhumed strike-slip faults is their lateral thermal gradient induced by the lateral juxtaposition of hot and cold levels of the crust controlling relevant properties of such fault zones, and thus the overall fault architecture (e.g., fault core, damage zone, shear lenses, fault rocks) and its thermal structure. These properties of the overall fault architecture include strength of fault rocks, permeability and porosity, the hydrological regime, as well as the nature and origin of circulating hydrothermal fluids.

© 2016 Published by Elsevier B.V.

Contents

1. Introduction .......................................................... 156
2. Review of deep structure of continental strike-slip faults ........................................ 156
3. Characterization of exhumed strike-slip faults ................................................ 156
   3.1. Temperature and depth-dependent changes ........................................ 158
   3.2. Processes related to strike-slip faults ........................................ 159
4. Strike-slip fault initiation at hot-to-cold contacts ........................................ 159
   4.1. Pluton-controlled strike-slip faults ........................................ 159
      4.1.1. Type I ........................................ 159
      4.1.2. Type II ........................................ 164
      4.1.3. Type III ........................................ 164
      4.1.4. Type IV ........................................ 164
   4.2. MCC-controlled strike-slip faults ........................................ 164
      4.2.1. Type A ........................................ 164
      4.2.2. Type B, pull-apart type MCCs ........................................ 164
   4.3. Transitional types between pluton- and MCC-controlled end-member strike-slip faults ........................................ 165

⁎ Corresponding author at: Dept. Geography and Geology, University of Salzburg, Hellbrunnerstr. 34, A-5020 Salzburg, Austria.
E-mail address: shuyun.cao@cug.edu.cn (S. Cao).

http://dx.doi.org/10.1016/j.earscirev.2016.09.010
0012-8252/© 2016 Published by Elsevier B.V.
1. Introduction

Formation of major exhumed strike-slip faults represents one of the most important dynamic processes significantly affecting the lithosphere-asthenosphere system. The exhumed strike-slip fault zones commonly preserved as long-standing zones of weakness in the Earth’s crust (Handy, 1989), and the manners in which they form are known to be of significant importance for earthquake mechanics (e.g., Wensoulsky, 1988; Stirling et al., 1996; Lyakhovsky et al., 2001; Carpenter et al., 2011) associated with surface displacements along their strike representing important global geological hazards (e.g., Rutter et al., 2001; Mooney and White, 2010). Near-surface rocks along faults are broken by cataclastic processes and have, therefore, no or a low strength. For a wide variety of practical purposes, for example, tunneling and underground excavation, exploration of hydrothermal ore deposits, geothermal energy and hydrocarbon exploitation (e.g., Aydin and Eyal, 2002; Sorkhabi and Tsuji, 2005; Weinberg et al., 2005; Bense et al., 2013; Wilson et al., 2013), the study and prediction of the anatomy of exhumed strike-slip fault zones is of particular interest. Due to high porosity, steep fault zones represent pathways for ascending and descending fluids and related alteration processes by fluid-rock interaction (e.g., Mittermeirger et al., 2009; Faulkner et al., 2010; Pei et al., 2015). This is also especially true of exhumed strike-slip faults in which mixed ductile and brittle fault rocks control the strength, porosity and permeability (Faulkner et al., 2010).

Continental-scale strike-slip fault zones are common tectonic features, particularly at convergent plate boundaries produced by oblique convergence and continental indentation (Storti et al., 2003). They are deep crustal expressions of the interaction between kinematic boundary conditions, rock rheology, and associated stress fields. During the last decades, many studies involved field observations, laboratory experiments, seismology, hydrogeology, and analytical and numerical modelling, have concentrated on describing and understanding the architecture, rock failure processes and mechanisms of the continental exhumed strike-slip zones (e.g., Allen, 1965; Katili, 1970; Fitch, 1972; Sibson, 1977; Scholz, 1980, 1989, 1990; Wise et al., 1984; Hannner, 1988; Sylvester, 1988; Woodcock and Schubert, 1994; Wintsch et al., 1995; Tikoff and de Saint Blanquat, 1997a; Läuer et al., 1997; Brown and Solar, 1998; Brown and Phillips, 1999; Teyssier and Tikoff, 1998; Paterson and Schmidt, 1999; Evans et al., 2000; Handy et al., 2001, 2005, 2007; Holdsworth et al., 2001; Storti et al., 2003; Rosenberg, 2004; Corti et al., 2005; Kim and Sanderson, 2006; Cunningham and Mann, 2007; Morrow et al., 2007; Finzi et al., 2009; Griffith et al., 2009; Widberley et al., 2008; Bistacchi et al., 2010; Molnar and Dayem, 2010; Frost et al., 2011; Keppler et al., 2013; Evans et al., 2014). However, the structural evolution and formation mechanisms of exhumed strike-slip faults and of their implications for lithospheric dynamics are still poorly understood. Detailed models of the potential initiation, and properties and architecture of orogen-scale exhumed strike-slip faults and how these relate to exhumation are rare.

In the following section, we review the state of exhumed continental-scale strike-slip fault systems, which represent a deep-crustal region across fault zones now exposed at the surface. We will discuss exhumed strike-slip faults, in which initially ductile structures are superimposed by brittle structures and their sequential formation relates to the exhumation of ductile fault rocks. The following controversial key issues are addressed: (1) Which relevant properties of such orogen-scale exhumed strike-slip fault zones with mixed ductile and brittle rocks can be deduced from fault sections? (2) What mechanisms exhume such faults? (3) How do orogen-scale strike-slip faults nucleate and initiate along rheologically weak zones? (4) How do thermal structure and fluids at the crustal level of fault change during various faulting and exhumation stages? The main emphasis of this review study is on the initiation of strike-slip faults at these end-member controlled tectonic settings. Many orogen-scale strike-slip faults initiate and further develop along rheologically weak linear zones at depth. These zones include (a) faults with abundant granite intrusions, (b) along lateral borders of hot metamorphic core complexes (MCCs), or (c) zones of rheological weakness due to ascending fluids or the presence of rheologically weak minerals such as talc or graphite.

2. Review of deep structure of continental strike-slip faults

Many continental strike-slip faults (Fig. 1) are observed to be weak compared with the surrounding rocks through evidence of geological, geophysical and laboratory measurements of frictional strength. The

![Fig. 1. Cartoon showing major continental intraplate strike-slip fault zone in the upper crust passing downwards into shear zones in the lower crust and lithospheric mantle. Schematic strength vs. depth profile or continental lithosphere shown to the left. Figure modified from Teyssier and Tikoff, 1998, Vauchez et al., 1998 and Storti et al., 2003.](image-url)
cause of this weakness is debated (e.g., Handy, 1989; Chester et al., 1993; Tikoff and Greene, 1997; Scholz, 2000; Hand et al., 2001, 2005, 2007; Faulkner and Rutter, 2001; Rutter et al., 2001; Holdsworth, 2004; Rosenberg, 2004; Rosenberg et al., 2007a, 2007b; Colletti et al., 2009; Niemeijer et al., 2010; Carpenter et al., 2011; Green et al., 2015). In early models, strike-slip systems were considered to expose a similar distinct structural level along their strike (Allen, 1965; Sibson, 1977). Strike-slip faults extend across the brittle crust with its seismicity through the brittle-ductile transition into the ductile lower crust (e.g., Sibson, 1977; Scholz, 1989) (Fig. 2). Numerous data indicate that the initial view of fault behavior in early models of strike-slip systems (e.g., Allen, 1965; Sibson, 1977; Wilcox et al., 1973; Mandl, 1988; Scholz, 1989) is oversimplified. These observations suggest that continental strike-slip faults generally develop in the middle and lower crust (ca. 10–20 km, depending on the geothermal gradient), and can be associated with a crustal detachment zone (e.g., Teyssier and Tikoff, 1998; Teyssier and Cruz, 2004) or near the base of the seismogenic zone (e.g., Molnar and Dayem, 2010; Meyer et al., 2014). A number of models indicate that subvertical brittle strike-slip faults in the upper brittle crust may potentially be rooted in subhorizontal ductile shear zones, implying a listric geometry of the fault plane in the lower crust and lithospheric upper mantle (Teyssier and Tikoff, 1998) (Fig. 1), which is constrained by geophysical methods (e.g., Mooney et al., 2007). It has been interpreted through deep seismic profiling that continental strike-slip faults transect the entire crust and roots in the upper mantle (e.g., Vauchez and Da Silva, 1992; Vauchez and Tommasi, 2003; Titus et al., 2007). Based on recent work, it is evident that continental strike-slip systems can occur in association with wide transrotational zones (e.g., Dickinson, 1996; Dewey et al., 1998; Doglioni et al., 2011) and wide zones of dispersed deformation mainly in transtensional to transpressional settings (e.g., Ring et al., 1992; Doglioni, 1995; Doglioni and Harbaglia, 1996; Dewey, 2002; Till et al., 2007a, 2007b). Recent studies initially based on analysis of field data integrated with numerical and analogue modelling experiments have been proven to be powerful tools in gaining insights into the evolution of continental-scale strike-slip fault systems (e.g., Handy et al., 2007; Wibberley et al., 2008; Frost et al., 2011; Corti et al., 2005; Keeller et al., 2013).

Crustal-scale intracontinental (transcurrent) strike-slip faults often develop at rheological boundaries between stiff and weak blocks. GPS data indicate a wide range from ca. 5 to 46 mm/yr of lateral displacement in major recently active strike-slip faults (e.g., Molnar and Dayem, 2010). Weak fault zones experience near-field uplift, strike-slip faulting in the borderlands, and strong fault-related strain to far-field deformation (e.g., Snoke et al., 1998; Cowgill et al., 2004; Dayem et al., 2009b; Titus et al., 2007). The distribution of near-field deformation may also be related to across-strike variations in lithospheric strength or thermal state (e.g., Griscom and Jachens, 1990; Wakabayashi et al., 2004; Molnar and Dayem, 2010; Chatzaras et al., 2015). Experimental rock deformation studies and new models have proposed explanations for the formation of the strike-slip fault zones. These examples include the strength of the lithosphere, shear heating, rocks rheology depends on composition as well as grain size, temperature, pressure, and localized presence of fluids (e.g., Handy et al., 2007; Colletti et al., 2009; Evans et al., 2000; Faulkner and Rutter, 2001; Bense et al., 2013). Thermally activated deformation mechanisms such as crystal plasticity and diffusional creep can lead to significant changes in rheological behavior and mechanical strength with depth (e.g., Sibson, 1977; Tullis and Yund, 1977; Scholz, 1989; Schmid and Handy, 1991; Wibberley, 1999; Rutter et al., 2001; Wibberley et al., 2008). Continental lithosphere exhibits lateral variations in mechanical

---

Fig. 2. Model showing a subvertical fault zone from surface to depth in continental crust at geothermal. (a) Profile showing the variation of structures of a crustal-scale fault zone from surface to depth (model as initially proposed by Sibson, 1977 and Handy et al., 2007). (b) Schematic strength vs. depth curve for the fault zone from pressure dependent brittle strength to temperature dependent plastic strength at depth.
In recent years, a number of publications have discussed the characterization of exhumed crustal-scale strike-slip faults and fault rocks relating deformation processes and damage zones on various scales (e.g., Chester, 1995; Evans et al., 2000; Ben-Zion and Sammis, 2003; Sibson, 2003; Di Toro and Pennacchioni, 2005; Moore and Rymer, 2007; Wibberley et al., 2008; Faulkner et al., 2010; Vaughan and Scarrow, 2003; Weinberg et al., 2004, 2009; Faulkner et al., 2010; Leever et al., 2011). The strike-slip faults have largely been passively exhumed, so that their present day surface structure can be viewed as representative of the structure and rheology of fault-related rocks change from mylonite, to fault breccias, fault gouge, kakirite, cataclasite and have pseudotachylyte at the transition between rocks, which are well known but mainly studied in hydrothermal ore deposits (e.g., Hedenquist and Lowenstern, 1994; Sibson, 2001, 2003). Since the inception of this fault zone model, however, this rather simple view has been modified to include transient mechanical behavior related to rate-dependent frictional sliding (e.g., Scholz, 1990, 1998). As the temperature increases with greater crustal depths, a frictional fault with pore-fluid pressure in the brittle regime, changes to a ductile shear zone with polymineralic plastic/viscous flow in the solid-state (Handy et al., 1999). Variation in physical-chemical conditions with depths results therefore in changes in the degree of strain localization in the uppermost crust and in the middle to lower crust (changes in the thickness of a shear zone, Fig. 2) (Sibson, 1977; Handy et al., 2007). Fault rocks formed at great depths during faulting are preserved in shear lenses with younger fabrics and are juxtaposed with later active fault cores, resulting in a “telescoped” fault zone structure.

In this fault zone model, we include plutonic and metamorphic rocks formed during faulting/shearing at depth preserved in shear lenses (Fig. 2). The brittle part of fault zones is distinguished by fault cores with anastomosing zones of cohesion-less rocks, mechanically stiff shear lenses and the damage zone surrounding the fault cores (e.g., Caine et al., 1996; Choi et al., 2016). The width ratio of the fault cores and damage zones has been used to classify fault zones because these influence the hydrological regime, technical properties (Caine et al., 1996), and ore mineralization (Sibson, 2001; Yardley and Baumgartner, 2007). Fault zones also generally provide pathways for the flow of hydrous and other fluids (e.g., CO2, nitrogen) (e.g., the San Andreas Fault Observatory at Depth – SAFOD) (e.g., Collins, 1979; Faulkner and Rutter, 1998; Clark and James, 2003; Becker et al., 2008; Vry et al., 2009; Faulkner et al., 2010). Fluid-rock interaction results in a wide range of altered rocks, which are well known but mainly studied in hydrothermal ore deposits (e.g., Hedenquist and Lowenstern, 1994; Sibson, 2001, 2003).
3.2. Processes related to strike-slip faults

Studies of exhumed fault zones have described the presence of rocks from great depth, as accommodating the bulk of fault zone displacement (e.g., Evans et al., 2000; Cole et al., 2007; Rosenberg and Schneider, 2008; Frost et al., 2009; Cao et al., 2011a, 2011b, 2011c). A number of processes have been discussed for the tectonic origin of orogen-scale strike-slip faults in mountain belts, including: (1) conjugate strike-slip faults triggered by indentation of a rigid block into a rheologically weak orogen belt, leading to lateral extrusion of the blocks (e.g., Tappinonier and Molnar, 1976; Ratschbacher et al., 1991); (2) tear faults at the boundaries of fold-thrust belts (e.g., Boyer and Elliott, 1982); (3) oblique-slip faults at the terminations of pull-aparts (Mann et al., 1983); and (4) the lateral boundary of retreating subduction zones (e.g., Royden, 1993; Doglioni et al., 1999a; 1999b; Wortel and Spakman, 2000) now termed STEP (Subduction-Transform Edge Propagator) faults (Govers and Wortel, 2005). Examples of exhumation in crustal-scale strike-slip fault zones have been recognized in both transtensional (e.g., the south Canadian Rockies and Tintina fault zone, Price, 2003) and transpressional (e.g., Alpine fault zone, New Zealand; Norris and Cooper, 2003; Little et al., 2005) tectonic settings. It is generally believed that rheological weakening mechanisms play an important role for fault localization as faults generally nucleate in the weakest zone (Schmid et al., 1996; Imber et al., 1997, 2008; Rosenberg, 2004; Hand et al., 2005, 2007; Dayem et al., 2009a; Molnar and Dayem, 2010; Yamasaki et al., 2014). Based on seismic behavior and geodetic surveys, Molnar and Dayem (2010) recently proposed that major intracontinental strike-slip faults develop along rheological boundaries between rigid, weak crust, and lithospheric mantle. Fossen and Rotevand (2016) suggest that the faults initiated as individual segments whose fault tips at some point interacted at the locations of the salient to form the curved fault segments. As a result, plate motion forces (ridge push and slab pull) are the main driving force to create strike-slip faults. This implies that the whole lithosphere is involved, and the initiation of strike-slip faults is likely at depth, where temperature-controlled rheological weakening mechanisms might play an essential role localizing future strike-slip faults.

As described in Chapter 3, exhumed strike-slip fault zones have a characteristic fault core, which is exposed and can be distinguished from the surrounding wider zone of damage (e.g., Evans et al., 2000) (Fig. 2). The fault core is the result of highly localized strain and intense shearing that accommodates the majority of the displacement within the fault zone. They generally consist of a number of recurring slip surfaces and fault rocks such as gouges, cataclasites, and mylonite (e.g., Sibson, 1977; Anderson et al., 1983; Chester et al., 1993; Bruhn et al., 1994; Caine et al., 1996; Evans et al., 2000; Faulkner et al., 2003; Wibberley and Shimamoto, 2003; Wibberley et al., 2008) (Fig. 3). The question of how the strike-slip fault zone structure varies as a function of depth, from the ductile lower crust up through the seismogenic crust to near the surface, remains unresolved. Several major processes are claimed to contribute to exhumation of the deeply seated rocks: (1) The most important exhumation process is likely oblique-slip motion with a partitioning of transpressive and transpressional deformation associated exhumation in an obliquely divergent setting with a significant but small vertical component. (e.g., Neubauer et al., 1995; Wang and Neubauer, 1998; Rosenberg, 2004; Cole et al., 2007; Frost et al., 2009) (Fig. 4A and B); (2) exhumation by superposition of earlier pure strike-slip fault (step 1 in Fig. 4C1) and subsequent normal faulting (step 1 in Fig. 4C2); (3) exhumation processes due to erosive flow or vertical extrusion (“pop-up”) within a positive flower or palm-shaped structure along a restraining bend by oblique reverse strike-slip faulting in a shortening setting (Sylvester, 1988; Woodcock and Rickards, 2003; Rosenberg, 2004) or at a small convergence angle (Leever et al., 2011) (Fig. 4D); (4) buoyancy driven exhumation of a low-density plutonometamorphic crust (Fig. 4E), encounters a rheologically weak layer or melt as is demonstrated by analogue transpressional wrenching experiments (Corti et al., 2003); (5) decompression and extension associated exhumation along the strike-slip fault system (Cao et al., 2011c); (6) unloading by tectonic unroofing along releasing bends or pull-aparts (Genser and Neubauer, 1988); and (7) Simple erosion of an orogeny may expose the deep structure of a strike-slip fault/shear zone (Handy, 1998). In this case, no superposition of ductile by brittle structures is needed to explain exhumation. However, it is still not well presented, which processes play the critical role in the exhumation of the deep-seated rocks in strike-slip fault zone systems.

4. Strike-slip fault initiation at hot-to-cold contacts

In this section, we focus on the tectonic settings where hot geological bodies are juxtaposed with cold ones, resulting in subsequent shear concentration along hot-to-cold contacts because of thermally-enhanced rheological weakening of the crust. Many examples, in which exhumed strike-slip faults or mylonitic strike-slip shear zones are in a close temporal and spatial relationship to hot plutos, magmatic gneiss domes, and/or metamorphic core complex rocks have been observed (e.g., Hollister and Crawford, 1986; Vandez and Da Silva, 1992; Hollister, 1993; Ingram and Hutton, 1994; Tommasi et al., 1994; Neubauer et al., 1995; Neves and Vaucluse, 1993; Berger et al., 1996; Neves et al., 1996; Vaucluse et al., 1997; Tikoff and de Saint Blanquat, 1997a; Brown and Solar, 1998; Pe-Piper et al., 1998; Wang and Neubauer, 1998; Paterson and Schmidt, 1999; Handy et al., 2001; Vaughan and Scarrow, 2003; Rosenberg, 2004; Weinberg et al., 2004; Cao et al., 2011a, 2011c). Examples of these faults include the Sumatra strike-slip fault (Fitch, 1972; Molnar and Dayem, 2010), and Ailao Shan-Red River strike-slip fault zone (Tappoinier and Molnar, 1976). Largely inactive major strike-slip faults sometimes contain plutonic and metamorphic rocks formed during faulting/shearing at depth preserved in shear lenses. These faulted plutonic or metamorphic rocks are then juxtaposed to non-metamorphic wall rocks. A basic requirement for the recognition of such strike-slip fault zones is the contemporaneity of associated structures such as the formation of granite intrusions, metamorphic (core) complexes and/or magmatic gneiss domes and potential formation of sedimentary basins at the surface. We attempt to classify the structures of strike-slip faults according to the overall geometry and using the contemporaneity criterion. Two end-member tectonic settings are distinguished, namely (1) plutons are associated with strike-slip faulting (Fig. 5A), and (2) hot metamorphic (core) complexes form one wall of the strike-slip/oblique-slip fault (Fig. 5B). The juxtaposition of hot, rheologically weak levels if crust to cool crust causes a strong horizontal temperature gradient to be developed. This contact guides the initiation of strike slip faulting by en echelon arrangement of uprising plutons (see Section 4.1). We also note gradual transitions between these two end-member settings. In both cases, the hot-to-cold contact laterally juxtaposes a rheological strong level with a very weak one.

4.1. Pluton-controlled strike-slip faults

Pre- and syntectonic plutons are emplaced preferentially within or at the margins of wide, regional-scale strike-slip fault zones, which intersect major lithological boundaries (Weinberg et al., 2004). Pluton-controlled strike-slip faults can be classified into several types. In all these cases, the fault core itself is strongly localization and weakened by the uprising magma. Basic types of the models are shown in Fig. 5A, and further examples in Figs. 6 to 8. Their locations on Earth are shown in Fig. 9. Significant data of such strike-slip dominated faults are compiled in Table 1.

4.1.1. Type I

Contemporaneously formed K-rich pre-kinematic plutonic rocks form directly related to strike-slip faults (Vaughan and Scarrow, 2003) (Fig. 6A). The K-magmatism results from melting of the metasomatized
base of the lithosphere (of a non-convecting mantle). The rising K-magma accumulations are triggered by transtension or transpression and weakening of the lithosphere. Uprising magma blobs are considered to coalesce and to form a strike-slip fault, which laterally propagates when appropriate stresses are involved (Vaughan and Scarrow, 2003). Examples include the Ailao Shan-Red River strike-slip faults (Cao et al., 2011a) (Fig. 6B, for location in Fig. 7), Elbe Zone of the Bohemian Massif (e.g., Wenzel et al., 2000) and Caledonian

Fig. 4. Some models of tectonic processes contributing to exhumation of the deep-seated rocks along major strike-slip fault zone systems. (A) Exhumation along transtensive oblique strike-slip faulting with constant orientation of displacement vector. The footwall exposes deeper levels of metamorphic core complex (MCC). (B) Exhumation along transpressive oblique strike-slip faulting with constant orientation of displacement vector. Hanging wall exposes the MCC. (C) Two-step exhumation by superposition of earlier pure strike-slip (step 1 in C1) and subsequent normal faulting (step 2 in C2). (D) Vertical exhumation by oblique reverse shear zone with ascending magma in a shortening setting (modified after Handy et al., 2001). (E) Self-exhumation by buoyancy driven due low density of plutonometamorphic crust (modified after Handy et al., 2001).

Fig. 5. Two end-members type models examples on initiation of hot-to-cold contacts on releasing oversteps of strike-slip faults. (A) Releasing overstep with syn-tectonic emplacement of granite reducing the strength of crust and forming a pluton-controlled strike-slip rheological boundaries with hot-to-cold contacts. (B) Releasing overstep delimiting a metamorphic core complex with hot-to-cold contacts.
Fig. 6. Some case examples of Types I and II on initiation of strike-slip faults at pluton-controlled rheological boundaries. (A) The plumbing system of K-magmatism forming at the base of the lithosphere initiates in zones of weakness within the lithosphere resulting in strain concentration (Vaughan and Scarrow, 2003). (B) Ailao Shan-Red River shear zone (modified from Leloup et al., 1995, Searle, 2013, Cao et al., 2011a and Liu et al., 2012). (C) Plutonic systems along island arc-parallel strike-slip zones partitioning oblique plate convergence into island-perpendicular convergence and island arc parallel strike-slip motion (Fitch, 1972; Molnar and Dayem, 2010). (D) Plutonic systems along island arc-parallel Sumatra Fault partitioning of oblique plate convergence into island-perpendicular convergence and island arc parallel strike-slip motion (modified after Fitch, 1972; Tikoff and de Saint Blanquat, 1997a; Acocella, 2014). For locations, see Fig. 7.

Fig. 7. Locations of some major exhumed pluton- and metamorphic core complex-controlled strike-slip faults in the world.
Fig. 8. Some case examples of Types III and IV on initiation of strike-slip faults at pluton-controlled rheological boundaries. (A) Slab-break-off magmatism (von Blanckenburg and Davies, 1995), the case of Periadriatic Fault (for locations, see Fig. 11). (B) Slab-break-off magmatism in the case of the Periadriatic Fault (modified after Rosenberg, 2004). (C) Defreggen-Antholz-Valles fault with the pull-apart-type Rieserferner tonalite body (modified from Stöckhert, 1985). For locations, see Fig. 7.

Fig. 9. Further case examples of Type IV on initiation of strike-slip faults at pluton-controlled rheological boundaries. (A) Shear zones related to post-collisional Variscan leucogranites in southern Brittany, France. Example is the South-Armorican shear zone (after Jégouzo, 1980). Pull-apart pluton of (B) is from southeastern Southern branch of the South Armorican shear zone. (C) Plutons along transtensional zones in island arcs, with the example of the Median Tectonic Line in the Japanese island arc (Taira, 2001). (D) Main Uralian strike-slip fault (modified from Hetzel and Glodny, 2002). For locations, see Fig. 7.
<table>
<thead>
<tr>
<th>Fault name</th>
<th>Location</th>
<th>Type</th>
<th>Length (km)</th>
<th>Shear sense</th>
<th>Timing of shearing</th>
<th>P-T conditions</th>
<th>Synkinematic</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alpine fault</td>
<td>New Zealand, Southern Island (transform fault between Pacific and Australian/Indian plates) North West Territories, Canada</td>
<td>MCC-controlled; plate boundary; orogen oblique in the Miocene; orogen-parallel Pluto-controlled; intracontinental orogen-oblique</td>
<td>Ca. 600</td>
<td>Dextral</td>
<td>Since ca. 36 Ma</td>
<td>Lower amphibolite facies T: 490–540 °C P: 5.4–6.8 k</td>
<td>Some pegmatitic dykes</td>
<td>Sibson et al., 1979; Grapes, 1995</td>
</tr>
<tr>
<td>Great Slave Lake shear zone</td>
<td></td>
<td></td>
<td>Ca. 1300</td>
<td>Dextral</td>
<td>2–1.9 Ga</td>
<td>Granulite to greenschist, 880 °C–680, 8.2 k to 4 k</td>
<td>Diorite, migmatisites, dykes</td>
<td>Hoffman, 1987; Hanmer, 1988</td>
</tr>
<tr>
<td>Ailaoshan–Red River fault</td>
<td>Boundary between Indochina and South China blocks</td>
<td></td>
<td>&gt;1000</td>
<td>Sinistral</td>
<td>Since ca. 31 Ma</td>
<td>Amphibolite facies T: 700 °C P: 5–6 k, retrogression to greenschist facies</td>
<td>Monzonitic granites, leucogranites</td>
<td>Leblond et al., 1995; Searle, 2006; Searle et al., 2010; Cao et al., 2011a</td>
</tr>
<tr>
<td>Dom Feliciano belt</td>
<td>Internal collision belt between Rio de la Plata &amp; Kalahari cratons (southern Brazil–Uruguay)</td>
<td></td>
<td>&gt;500</td>
<td>Sinistral</td>
<td>SINAPAfrican</td>
<td>Prograde metamorphism from ca. 550 °C to 675 °C; ca. 425 °C at ca. 3 k</td>
<td>Diorite-adelamellite; sodic-potassic to peralkaline granites; leucogranites</td>
<td>Jupp et al., 1994</td>
</tr>
<tr>
<td>South American shear zone</td>
<td>Separates Central and South American domains (Southern Brittany, France)</td>
<td></td>
<td>&gt;300</td>
<td>Dextral</td>
<td>Carboniforous</td>
<td>Triplet point And-Sill-Ky (ca. 450 °C, 4.5 k) to greenschist</td>
<td>Leucogranites</td>
<td>Jégouzo, 1980; Faure et al., 2008; Lemarchand et al., 2012</td>
</tr>
<tr>
<td>Coimbra-Badajoz-Cordoba shear zone</td>
<td>Central Portugal and Spain (Southern Brittany, France)</td>
<td></td>
<td>&gt;400</td>
<td>Sinistral</td>
<td>Upper Carboniferous faces</td>
<td>Granite-ap amphibolite-greenschist facies</td>
<td>Two mica-leucogranite leucogranites</td>
<td>Burg et al., 1981</td>
</tr>
<tr>
<td>Serra da Freita shear zone</td>
<td>Central Portugal</td>
<td></td>
<td>&lt;70</td>
<td>Sinistral</td>
<td>SINAPAfrican</td>
<td>Amphibolite facies (sillimanite)</td>
<td>Hutton and Reevy, 1992</td>
<td></td>
</tr>
<tr>
<td>Sumatra fault</td>
<td>Indonesia</td>
<td></td>
<td>Ca. 1900</td>
<td>Dextral</td>
<td>SINAPAfrican</td>
<td>SINAPAfrican</td>
<td>Granitoids</td>
<td>Fitch, 1972; Alvarado et al., 2011; Sieh and Natwidi, 2000</td>
</tr>
<tr>
<td>Periadiatic fault</td>
<td>Separates the Southalpine unit from Austroalpine/Penninic units</td>
<td></td>
<td>Ca. 700</td>
<td>Dextral</td>
<td>SINAPAfrican</td>
<td>SINAPAfrican</td>
<td>Granitoids</td>
<td>Stöcklin et al., 2011; Handy et al., 2005; Rosenberg, 2004; Fodor et al., 1998</td>
</tr>
<tr>
<td>Median tectonic line</td>
<td>SW Japan</td>
<td></td>
<td>Ca. 300</td>
<td>Dextral</td>
<td>Cretaceous</td>
<td>Amphibolite to greenschist facies</td>
<td>Granioids</td>
<td>Taira, 2001; Jeffries et al., 2006; Wübbeler and Shimamoto, 2003</td>
</tr>
<tr>
<td>Defreggen-Antholz-Valles fault</td>
<td>Separates Oligocene Tauern metamorphic terrain from Unmetamorphic terrain</td>
<td></td>
<td>Ca. 100</td>
<td>Sinistral</td>
<td>SINAPAfrican</td>
<td>Greenschist facies</td>
<td>Tonalite</td>
<td>Stöckhert, 1985; Kleinschmidt, 1987; Steenken et al., 2000; Wagner et al., 2006</td>
</tr>
<tr>
<td>Karakoram fault</td>
<td>Across India and Asia in the Himalaya region</td>
<td></td>
<td>Ca. 800</td>
<td>Dextral</td>
<td>SINAPAfrican</td>
<td>Amphibolite-facies to greenschist-facies</td>
<td>Granitoids</td>
<td>Weinberg et al., 2009; Boutonnet et al., 2012; Wallis et al., 2014</td>
</tr>
<tr>
<td>Northern Borborema Province shear zones</td>
<td>European Eastern Alps; borders the Oligocene-Miocene metamorphic Penninic units of the Tauern window Borborema province of Northern Brazil</td>
<td></td>
<td>Ca. 450</td>
<td>Sinistral</td>
<td>SINAPAfrican</td>
<td>Upper amphibolite-facies to greenschist-facies</td>
<td>Granitoids</td>
<td>Neves et al., 1996</td>
</tr>
<tr>
<td>Salzach-Ennstal–Mariazell–Puchberg fault system (SEMP)</td>
<td>European Eastern Alps; borders the Oligocene-Miocene metamorphic Penninic units of the Tauern window Borborema province of Northern Brazil</td>
<td></td>
<td>Ca. 30–32 Ma</td>
<td>Dextral</td>
<td>SINAPAfrican</td>
<td>Upper to lower greenschist facies conditions</td>
<td>Granitoids</td>
<td>Pereson and Deckard, 1997; Wang and Neubauer, 1998; Cole et al., 2007; Glodny et al., 2005; Frost et al., 2011; Schneider et al., 2013</td>
</tr>
<tr>
<td>Gleinalm shear zone</td>
<td>European Eastern Alps</td>
<td></td>
<td>Ca. 80</td>
<td>Sinistral</td>
<td>SINAPAfrican</td>
<td>Amphibolite to greenschist facies</td>
<td>Neubauer et al., 1995</td>
<td></td>
</tr>
<tr>
<td>Chakragil–Ghez fault</td>
<td></td>
<td></td>
<td>&gt;200</td>
<td>Dextral</td>
<td>SINAPAfrican</td>
<td>Granitoids</td>
<td>Neubauer et al., 1995</td>
<td></td>
</tr>
<tr>
<td>Neybar–Chatak shear zone</td>
<td></td>
<td></td>
<td>Ca. 100</td>
<td>Sinistral</td>
<td>SINAPAfrican</td>
<td>Amphibolite to greenschist facies</td>
<td>Neves et al., 2013; Cao et al., 2013a, 2013b</td>
<td></td>
</tr>
<tr>
<td>Tancheng–Lujiang (Tan–Lu) fault</td>
<td>Separates North from South China block</td>
<td></td>
<td>Ca. 2400</td>
<td>Sinistral</td>
<td>SINAPAfrican</td>
<td>Amphibolite to greenschist facies</td>
<td>Ratschbacher et al., 2000; Zhu et al., 2005</td>
<td></td>
</tr>
</tbody>
</table>

References:
- Hoffman, 1987
- Hanmer, 1988
- Leap et al., 1995
- Searle, 2006
- Searle et al., 2010
- Cao et al., 2011a
- Hoffman, 1987
- Hanmer, 1988
- Leap et al., 1995
- Searle, 2006
- Searle et al., 2010
- Cao et al., 2011a
- Hoffman, 1987
- Hanmer, 1988
- Leap et al., 1995
- Searle, 2006
- Searle et al., 2010
- Cao et al., 2011a
- Hoffman, 1987
- Hanmer, 1988
- Leap et al., 1995
- Searle, 2006
- Searle et al., 2010
- Cao et al., 2011a
- Hoffman, 1987
- Hanmer, 1988
- Leap et al., 1995
- Searle, 2006
- Searle et al., 2010
- Cao et al., 2011a
- Hoffman, 1987
- Hanmer, 1988
- Leap et al., 1995
- Searle, 2006
- Searle et al., 2010
shoshonitic magmatic belt of the northern British Isles (e.g., Vaughan, 1996).

4.1.2. Type II

Plutonic systems occur along island arc-parallel strike-slip fault zones or strike-slip faulting occurs along the axis of a syn-magmatic tectonic arc in a transpressional setting (Fig. 6C). Particularly in this case, the initiation of strike-slip by en echelon arrangement of uprising plutons plays a critical role for localization of the strike-slip fault zone. In such settings, strike-slip fault zones are often accompanied by volcanic centers, e.g., the Sumatra strike-slip fault on the Indonesian island (e.g., Fitch, 1972; McCaffrey, 1992, 2009; Acocella, 2014) (Fig. 6D, for location in Fig. 7), the Rosy Finch strike-slip fault zone in the eastern central Sierra Nevada (e.g., Tikoff and de Saint Blanquat, 1997b), the Philippine fault (Fitch, 1972), the El Salvador shear zone (Alvarado et al., 2011) and the western Idaho shear zone (Giorgis et al., 2008). In this case, the strike-slip zone is responsible for partitioning of oblique plate convergence into island-perpendicular convergence and island arc parallel strike-slip motion (e.g., Fitch, 1972; McCaffrey, 2009; Molnar and Dayem, 2010). The magma causes the fault to be localized by thermal weakening of the crust and even of the lithospheric mantle in the latter case when mantle-derived magmas are involved (e.g., Beck, 1983; Tikoff and de Saint Blanquat, 1997b).

4.1.3. Type III

Pluton development is often associated with strike-slip fault zones above slab break-off areas formed during episodes of oblique continent-continent collision (Fig. 8A). The lithosphere is predominantly affected by the formation of a slab window resulting in increased heating of the base of the upper lithosphere (e.g., Davies and von Blanckenburg, 1995; von Blanckenburg and Davies, 1995; Wortel and Spakman, 2000; Dayem et al., 2009a). Also in this case, the uprising magmas are weakening the mantle and the lower crust, which results in strain localization during the uprising along the uprising plutons. In some cases, strongly deformed pre- and syn-magmatic plutonic rocks occur with the strike-slip faulting, as well as typical bodies of sigmoid shape converging with a thin tail towards the strike-slip fault. Examples include the eastern part of the Periadriatic Fault (von Blanckenburg and Davies, 1995; Fodor et al., 1998; von Blanckenburg et al., 1998; Rosenberg, 2004; Fig. 8B, C) and the South Armorican shear zone (Jégouzo, 1980) (Fig. 9A), with pull-apart type plutons juxtaposed with much colder country rocks (Fig. 9B). Leucogranitic S-type plutons occur in the post-collisional strike-slip shear zone during wrenching in the late stages of oblique continent-continent collision (e.g., Faure and Pons, 1991; Searle, 2013). The mechanisms creating such post-collisional plutons along strike-slip systems are: (1) overall wrenching between the involved continental microplates and (2) a heating mechanism by delamination of the mantle lithosphere and heat input and/or shear heating. Well-known examples for such post-collisional shear-zone-related leucogranites are the two branches of the South Armorican shear zone (Jégouzo, 1980; Faure et al., 2008; Tartèse et al., 2012; Lemarchand et al., 2012; Fig. 9A).

4.1.4. Type IV

Major plutons occur along transtensional zones in island arcs or continental arcs although it remains often unclear whether the faults were initiated by pre-existing plutons or the strike-slip faults allowed for the ascent of magma. This type is often associated with coeval half-graben-type sedimentary basins along the ductile oblique-slip, transtensional fault. Examples include the Median Tectonic Line in the Japanese island arc in Japan (e.g., Taira, 2001; Fig. 9C) and the Ancestral Cascades arc in the southwestern United States (e.g., Busby et al., 2012).

On a smaller scale, pull-apart type plutons potentially rise up along releasing oversteps of strike-slip fault systems (Román-Berdiel et al., 1997). Such settings could also be considered as a subtype of the island arc-parallel strike-slip faults (Type II) or leucogranitic strike-slip zones (included in Type III). The strike-slip faulting at releasing oversteps represents a potential mechanism to create space for the intrusion at shallow crustal level. In such pull-apart-type plutons, margins are generally deformed in a ductile manner. The uprising bodies are part of a major plumbing system and intrude into active strike-slip faults, enhancing and focusing motion along such strike-slip fault zones where low strength zones are juxtaposed with high-strength country rocks. A mutual feedback loop of the intrusion of plutons with strike-slip faulting results in weakening of the crust and potentially of the whole lithosphere. Examples for pull-apart type plutonic bodies include the Rieserferner pluton in the Eastern Alps, which intruded into the Defreggen-Antholz-Valles Fault (Stockhert, 1985; Stöckli et al., 2001; Fig. 9D) and those along the South Armorican shear zone (Román-Berdiel et al., 1997) (Fig. 9A).

4.2. MCC-controlled strike-slip faults

In a horizontal section, hot MCCs are often juxtaposed with cold country rocks along subvertical or dipping detachment faults forming a major rheological contrast. In some cases, strike-slip faults are involved in confining structures of MCCs and the oblique-slip faults often connect with extensional detachments (e.g., Genser and Neubauer, 1988; Neubauer et al., 1995; Meyer et al., 2014) in a large-scale releasing overstep or termination of a crustal-scale strike-slip fault. It is worth noting that such strike-faults have properties of stretching faults in which one wall on the side of the MCC, behaves in a ductile manner (Means, 1989). A pre-condition for formation of this type of strike-slip faults is progressive metamorphism prior to initiation of strike-slip or oblique-slip faulting, whereas the juxtaposition of hot-to-cold contacts occurs during exhumation. For initiation, an instability condition for the localization of shearing along a strong vs. weak boundary is needed. The instability could represent the rheological contrast (that is a strong vs. a weak lithology, e.g., feldspar-dominated vs. calcite-dominated rocks within the crust) or simply the presence of rheologically very weak minerals such as talc or graphite (e.g., Lockner et al., 2011; Oohashi et al., 2013) along a pre-existing high-angle normal fault. Metamorphic (core) complexes along transpressional and transtensional strike-slip faults can be classified into two systems shown in Figs. 10, 11 & 12; further examples are presented in Fig. 13, and significant data are compiled in Table 1.

4.2.1. Type A

Transpressive or transtensive strike-slip zones with reverse or high-angle normal faults represent the boundary wall of the hot MCC. The metamorphic complex is tilted away from the fault, and the rocks with the highest P-T conditions are generally exposed very close to the fault itself or the fault axis whereas the metamorphic P-T conditions gradually decrease away from the confining strike-slip or oblique-slip fault. Coeval sedimentary basins are usually not associated with this type of fault. Such structures exhume the metamorphic complex mainly on one side of a strike-slip zone by either oblique-slip faulting or a superposed strike-slip and reverse motion. Examples include the Alpine Fault in New Zealand, where the existence of a footwall oblique ramp was postulated (Rowland and Sibson, 2004; Little et al., 2005; Chatzaras et al., 2013) (Fig. 10A). Another example is the Lepontine metamorphic dome along the western part with the Periadriatic Fault in the Alps (Fig. 10B), which juxtaposes the upper amphibolite-grade Lepontine rocks in the footwall to the nearly unmetamorphic South Alpine unit along the Periadriatic fault and less metamorphosed rocks above ductile low-angle normal faults including the Simplon and Forcola normal faults (e.g., Ellert et al., 2013).

4.2.2. Type B, pull-apart type MCCs

Releasing overstep of a strike-slip fault system occurs along pull-apart type MCCs. Such structures seem to be quite common and have been described several times (see below). Such structures include...
oversteps of strike-slip fault systems associated with ductile low-angle normal faults along upper margins. The metamorphism often reaches lower amphibolite-grade metamorphism (e.g., Yin, 2004). The interior of a MCC is often folded to form a dome or an antiform structure with folded metamorphic isogrades with the highest P-T conditions in the center of the dome. Coeval sedimentary basins can be associated with confining strike-slip faults or are on.

the hanging wall unit above the ductile low-angle normal faults. In these cases, the basins are either half-graben or collapse basins or transtension-edge basins. Examples of pull-apart type MCCs include the Cenozoic Taiern window MCC (Genser and Neubauer, 1988; Neubauer et al., 1999; Fig. 11C), the Cenozoic Eastern Pamir dome (Fig. 12A; Cao et al., 2013a, 2013b; Rasmus et al., 2013), the Panafriean Metabtq and Sibai domes in the Eastern Desert of Egypt (Fritz et al., 2002; Fig. 12B), the Panafriean Qazaz dome on the Arabian peninsula (Meyer et al., 2014; Fig. 12C), the Montagne Noire MCC of the European Varisicids (Echtler and Malavieille, 1990; Fig. 12D), the Upper Cretaceous Glenualm dome in the Eastern Alps (Neubauer et al., 1995; Fig. 12E), and the Karakoram metamorphic complex in the Himalaya region (Wallis et al., 2014; Fig. 12F). All these locations are presented in Fig. 7.

4.3. Transitional types between pluton- and MCC-controlled end-member strike-slip faults

The two types have gradual transitions from pluton-controlled strike-slip fault systems to MCC-controlled endmembers. This is particularly the case when syn-tectonic plutons intrude into the strike-slip fault zone and are deformed in the process. The Ailao Shan-Red River shear zone is a good example of such a superposition of MCC and pluton-controlled strike-slip zones (e.g., Fig. 6B, bordering the Cenozoic Diancang Shan MCC; Cao et al., 2011a, 2011b; Liu et al., 2012) in Asia. Other major strike-slip faults border metamorphic (core) complex associated with plutons within orogens. Examples include the Karakoram strike-slip fault (Boutonnet et al., 2012) (Fig. 12A).

Studies of plutonic and metamorphic rocks within strike-slip fault systems found that at least part of their exhumation is synchronous with strike-slip faulting (Nelson et al., 1996; Sippl et al., 2013). It is thus proposed that crustal-scale strike-slip faults may localize and nucleate due to thermal anomalies associated with preexisting granitic plutons/metamorphic complexes. This leads to strain localization (local reduction of the crustal strength) and preferential weakening, in the heated portions of the ductile lower crust.

4.4. Oroclines, syntaxes and exhumed strike-slip shear zones

An interesting question now arises from the observation that, in a number of cases, exhumation of major strike-slip faults is related to oroclinal syntaxes at the lateral tip of major indenting continental plates during collisional orogeny. Examples include: Pamir, Alaska and the Tauern window.

In the northwestern Himalayan syntaxis in front of the Indian plate, the Pamir exposes several high-grade metamorphic gneiss domes intruded by many granitoids (e.g., Schmidt et al., 2011; Stearns et al., 2013a, 2013b; Stübner et al., 2013; Fig. 13A). The gneiss domes are rooted in a rheologically hot and weak zone (Sippl et al., 2013). At their eastern termination, the Pamir gneiss domes are related to strike-slip faults, e.g., the Kongur Shan, Karakorum and Chakr-Gez strike-slip faults as discussed above. The granites seem to be smeared out from the Pamir and Kongurshan gneiss domes. Consequently, a genetic relationship between metamorphic gneiss domes, pluton, and MCC controlled
Fig. 11. Type B examples of MCC-controlled strike-slip faults from the Alps. (A) Simplified structural map of the Eastern Alps with locations of major exhumed strike-slip faults (modified after Pfiffner, 2014; or location, see Fig. 7). The late-stage, Late Oligocene-Miocene structure of the Alps is characterized by indentation of the rigid Adriatic microplate, decollement of previously subducted European continental crust along the Sub-Tauern ramp. (B) Crustal-scale TRANSALP cross-section showing the Tauern metamorphic core complex, the location of the sub-Tauern ramp, the antiformal doming of detached crust in the Tauern MCC and the involvement of major orogen-parallel strike-slip faults involved in the exhumation of the Tauern MCC belts (modified after TRANSALP Working Group et al., 2002). (C) Tauern MCC and SEMP fault as well as the Periadriatic fault decorated by syn-tectonic tonalites (modified from Genser and Neubauer, 1988; Neubauer et al., 1999) (for tonalites, see also Fig. 8B). (D) A model showing, the formation of the Tauern MCC, due to the oblique post collisional plate convergence and initiation of exhumation along a hot-to-cold contact along an oblique-slip fault.
exhumed strike-slip faults exist. In these cases, the strike-slip faults are characterized by amphibolite facies-grade metamorphic zones within the fault to nearly non-metamorphic country rocks.

A similar structural relationship can be postulated for southeastern Alaska, where the Eocene amphibolite facies-grade Chugach metamorphic complex was exhumed in the orocline above the subducting Kula plate. At the edge, the Border Range Fault, which is a ductile strike-slip fault, was activated (Gasser et al., 2011; Fig. 13B). The Border Range Fault juxtaposes the Chugach metamorphic complex to the largely non-metamorphic Wrangelia terrane and some granitoids intruded into this terrane. Interestingly, the area contains mesothermal gold mineralization, which formed during exhumation. Exhumation of the Eocene metamorphic complex is contemporaneous with fault activity (Gasser et al., 2011).

A third example is the Tauern MCC and SEMP fault as well as the Periadriatic fault evolved as a late-stage, Late Oligocene-Miocene orogen-parallel MCC and strike-slip fault zones in front of the rigid Adriatic indenter, which moved into the mechanically weak orogen (Ratschbacher et al., 1989, 1991). For the Tauern MCC see Fig. 11A. A seismic section (TRANSALP section, Fig. 11B) reveals that the Tauern MCC formation is characterized as a result of the decollement of a piece of previously subducted European continental crust and subsequent shortening and initiation of vertical extrusion along a hot-to-cold contact along an oblique-slip fault above the sub-Tauern ramp (TRANSALP Working Group et al., 2002) during oblique collision of the Adriatic plate (in a hanging wall position and the European plate in the footwall). This fault was connected with a blind thrust at depth, resulting in doming and folding of the metamorphic isograds (Lammerer and Weger, 1998; Neubauer et al., 1999; Pfiffner, 2014). The further development is driven by tectonic unroofing and lateral extrusion as discussed above. Along the western and eastern Tauern MCC, ductile low-angle normal faults accommodated orogen parallel stretching. Recently, Keil and Neubauer (2015) postulated east-directed raft tectonics above the eastern Katschberg ductile low-angle normal fault as a result of the topographic gradient due to antiformal doming of the Tauern window. The final shape of the Tauern MCC arose during late-stage bending of the mechanically weak interior by indentation through the brittle upper plate (Neubauer, 2005; Schmid et al., 2013).

5. Faulting mechanisms of an exhumed strike-slip fault

5.1. Dynamic weakening mechanisms

As discussed above, many continental strike-slip faults are observed to be weak compared with the surrounding country rock by evidence of geological, geophysical and laboratory measurements of frictional strength, but the cause of this weakness is debated (Chester et al., 1993; Scholz, 2000; Faulkner and Rutter, 2001; Holdsworth, 2004,
Holdsworth et al., 2010; Collettini et al., 2009; Niemeijer and Spiers, 2007; Niemeijer et al., 2010; Carpenter et al., 2011; Green et al., 2015). Recently, Rice (2012) proposed that major faults are statically strong but are dynamically weakening during seismic slip.

Aside from syntectonic plutons and MCCs, a number of other processes could potentially be responsible for dynamic weakening of exhumed strike-slip faults (e.g., Gratier and Gueydan, 2007; Rice and Cocco, 2007; Tullis et al., 2007; Tagami, 2012), (2) metamorphic reaction weakening by formation of extremely rheologically weak low-friction minerals such as talc, graphite, and clay minerals (e.g., Chester et al., 1993; Wibberley, 1999; Moore and Rymer, 2007; Collettini et al., 2009; Lockner et al., 2011; Kuo et al., 2014), (3) fluid migration by release of fluids in front of uprising magma, metamorphic dehydration reactions at depth and release of fluids from a subduction zone (e.g., Sibson, 2001), and (4) phyllosilicate-bearing fault gouges which develop velocity weakening behavior at rapid slip rates (e.g., Jefferies et al., 2006; Niemeijer and Spiers, 2007). The presence of rheologically weak minerals such as clay minerals in fault gouges in uppermost crustal levels (e.g., Hickman et al., 2008) is generally observed in quartzofeldspathic lithologies resulting from fluid-host rock interaction. The presence of

Fig. 13. Oroclinal bending, exhumation of a metamorphic core complex and associated strike-slip faults. (A) Pamir with gneiss domes and associated strike-slip faults such as the Chakragil-Ghel and Karakoram exhumed strike-slip faults (combined after Sippl et al., 2013 and Stübner et al., 2013). (B) Eocene Alaska with the Border Range strike-slip exposing the Chugach metamorphic complex (after Gasser et al., 2011).
graphite and talc and other rheologically weak minerals at shallow levels along major fault zones is used to explain the aseismic slip, e.g., along the creeping segment of the San Andreas Fault (Titus et al., 2006, 2011). The varying interplay factors between faulting, fluid migration, and hydrous clay mineral transformations along fault zones are suggested to constitute important weakening mechanisms. However, Collettini et al. (2009) suggested that dynamic weakening mechanisms can explain some observations (Imber et al., 1997). For example, creep and aseismic slip are thought to occur on weak faults, and quasi-static weakening mechanisms are required to initiate frictional slip on mis-oriented faults, at high angles to the tectonic stress field (Stewart et al., 2000). In most strike-slip fault zones, it is likely that more than one mechanism operates in any one part of the faulting at any given time. Moreover, the maintenance of high fluid pressures requires specialized conditions and weak mineral phases not to be present in sufficient abundance to satisfy weak fault models, so weak faults remain largely unexplained (Collettini et al., 2009).

5.2. Role of thermal structure of strike-slip faults

Recent experiments and observations from rocks and rock-like materials show that the thermal regime of fault zones mainly depends on three parameters. These are the regional geothermal structure across the fault zone and background thermal history, frictional heating of wall rocks by fault motions, and heating by advection of hot fluids within the fault core and adjacent damage zone (e.g., Handy et al., 2007; Tagami, 2012 and references therein). The thermal structure of strike-slip faults and shear zones seems to be linked to deformation mechanisms occurring in the hot or cold lithosphere (Fig. 14A, B) or to small-scale mechanisms, i.e. dynamic recrystallization and grain size reduction of rock-forming minerals (e.g., Yamasaki, 2004; Tagami, 2012). In this case, exhumed high-grade MCC or plutons associated with major strike-slip faults are of particular interest.

Both the thermal and mechanical effects related to the presence of magma and a hot MCC strongly weaken the lithosphere. This effect may have important implications for the model of continental-scale strike-slip faults as proposed in Fig. 14A and B. At a crustal scale, magmatic intrusions may significantly affect the strike-slip fault patterns, also favoring the development of the MCC structure and evolution (e.g., Corti et al., 2003). Scholz (1980) reviewed a number of cases of shear heating revealed by metamorphic aureoles around a fault zone. The steady-state numerical models assume and calculate shear heating in the whole lithosphere for both brittle and ductile deformation regimes (e.g. Leloup et al., 1999; Tagami, 2012). These cases all occur in ductile-deforming terrains where the strong temperature sensitivity of the rock strength results in thermal softening which provides a buffer for shear heating (for models, see Brun and Cobbold, 1980; Fleitout and Froidevaux, 1980). If the high-slip velocities are involved, the low thermal conductivity of rocks and the transient heating during rapid co-seismic slip essentially occurs without conduction, if so, and the fault is thin enough, very high temperatures may be reached locally (Rice, 2012). The various dynamic weakening processes include flash heating at highly stressed frictional micro-contacts and rapid slip reduces friction (e.g., Rice, 1999, 2012). Emplacement and rise of magmas and/or hot fluids in the shear zone will further enhance the temperature increase to shallower parts of the fault zone (e.g., Yardley and Baumgartner, 2007; Rice, 2012).

5.3. Impact of hot fluids in strike-slip faults

The interplay between the different fluid sources with host rocks has a strong influence on the thermal structure, as well as metamorphic and deformation processes of the strike-slip faults, with cooling effects by descending cold fluids of meteoric origin and heating by ascending hot fluids of magmatic and metamorphic origin (e.g., Faulkner and Rutter, 2001; Yardley and Baumgartner, 2007) (Fig. 14). Hot fluids are derived from ascending magma and metamorphic devolatilization, or may even rise up from subduction zones, e.g. caused by deserpentinization of subducted oceanic lithosphere (e.g., Saffer and Tobin, 2011; Guillot and Hattori, 2013). Reactions of rocks with hydrous fluids cause softening of metamorphic fabrics by means of phyllosilicate formation, lowering friction and resulting in phyllonites. These processes lead to the high permeability of fault zones (e.g., Wibberley et al., 2008; Fig. 14). Models invoking fluid pressure-controlled fault weakening and earthquake slip instability (e.g., Byerlee, 1990, 1993; Rice, 1992). High-temperature fluid-rock interactions occur widely during seismic slip and the geochemical characteristics of fault rocks are useful indicators of such coseismic events. Also seismic pumping of fluids is a common phenomenon (e.g., Sibson, 2001). All these effects are likely the reasons for strain localization along the strike-slip faults.

The hot fluids are highly reactive, dissolving elements from country rocks and resulting in many effects including phyllosilicate formation by

Fig. 14. Predictive models on important thermal and structural properties of crustal-scale strike faults. (A) Exhumed strike-slip fault containing a syntectonic pluton (example: eastern Periadriatic Fault) showing a hot fault core contrasting colder borders on both sides. Near surface, descending meteoric water may cool the fault. (B) Exhumed strike-slip fault bordering a metamorphic core complex showing a hot MCC juxtaposed to a cold block creating an asymmetric thermal gradient. The thermal structure of fault zones could be checked by thermochronological methods, e. g. zircon (U-Th)/He geochronology.
6. Initiation of major strike-slip faults: chicken or egg-first?

A major question is how and where a major strike-slip fault is initiated. Many crustal-scale strike-slip faults appear to form persistent zones of weakness that fundamentally influence the distribution, architecture and kinematic patterns of crustal-scale deformation and associated geological processes (e.g., Collettini et al., 2009 and references therein). Common in all models of strike-slip initiation is an important contrast of weak and strong rheologies. Models assume an important rheological contrast between rheologically weak and strong lithologies, for example margins of a stief craton and juxtaposed mobile belts (Weinberg et al., 2004; Furlong et al., 2007; Molnar and Dayem, 2010).

Some models assume weakening of the lithosphere due to the rise of magma. Magmatic weakening of the lithosphere has received particular attention in the context of continental faulting. The spatiotemporal relationship and compatible rates of processes between pre-kinematic or contemporaneous granitic plutons and crustal-scale strike-slip faulting have been fully documented worldwide in a number of cases (e.g., Guinebretier, 1987; Glazner, 1991; Hutton et al., 1990; D’lemos et al., 1992; Hutton and Reavy, 1992; McCaffrey, 1992; Tikoff and Teyssier, 1994; Brown, 1994; Davidson et al., 1994; Neves and Vauchez, 1995; Tikoff and de Saint Blanquat, 1997a; Tommasi et al., 1994; Neves et al., 1996; Vauchez et al., 1997; Brown and Solar, 1998; St. Blanquat et al., 1998; Handy et al., 2001; Alvarado et al., 2011; Rosenberg, 2004; Rosenberg et al., 2007b; Cao et al., 2011a; Brown, 2013). Since emplacement, crystallization and cooling of pre- and syntectonic granitic plutons may represent relatively rapid geological processes (e.g., Paterson and Tobisch, 1992; Karlstrom et al., 1993; Bodorkos et al., 2000; Petford et al., 2000; St. Blanquat et al., 2011; Zibra et al., 2012) they may also record short- or long-term geological events related to crustal melting and deformation of the continental crust. However, it is a matter of debate whether magma emplacements have previously localized and weakened the crust/lithosphere and triggered nucleation of shear zones/fault zones (e.g., Paterson and Fowler, 1993; Neves and Vauchez, 1995; Neves et al., 1996; Paterson and Schmidt, 1999; Vaughan and Scarrow, 2003, Weinberg et al., 2004; Rutter and Mecklenburg, 2005), or whether pre-existing faults or shear zones represent the preferred pathways for the ascension and emplacement of granitic magmas (e.g., Hutton et al., 1990; D’lemos et al., 1992; Holm, 1995; Pe-Piper et al., 1998; Sawyer, 1999; Brown and Solar, 1998, Brown, 2013). This leads to a chicken and egg debate over the chronological relationship between pluton-controlled exhumed strike-slip faults with pre- and/or syn-tectonic magmatism.

One solution of the dilemma is that granitic magma ascending and deformation are linked in a positive feedback loop (Brown and Solar, 1998) such that magma-enhanced strain localization leads to further magma segregation and crustal differentiation (Brown, 1994; Handy et al., 2001). As discussed in Chapter 3.2, the initiation of strike-slip fault zones is suggested to result from plate collision (e.g., escape tectonics; Tapponnier et al., 1982) or oblique collision mechanical modes, which induce lithospheric deformation (Fitch, 1972). The oblique subduction of the oceanic plate partitions in causes orthogonal shortening at plate margins and orogen-parallel strike-slip motion in the interior of the fossil and active island arc systems (Fitch, 1972; Molnar and Dayem, 2010). Oblique convergence along a subduction margin commonly results in localization of strike-slip deformation in the thermally weakened crust of the magmatic arc. The outstanding examples of this behavior include the Main Uralian strike-slip fault (Hitzel and Clodny, 2002), the Great Sumatra Fault (Fitch, 1972; Alvarado et al., 2011) and in the Japanese island arc (e.g. Taira, 2001). The Great Sumatra fault is accompanied by volcanoes, implying coeval magma chambers/plutons at depth as usually only 10–15% of the erupted magma, and pull-apart structures related to releasing step-overs of the Sumatran Fault (Bellier and Sébrier, 1994; Acocella, 2014). Von Blanckenburg and Davies (1995) proposed rheological weakening of the crust by slab break-off magmatism, which gives rise to future orogen-parallel strike-slip accommodation (Ratschbacher et al., 1991). Vaughan and Scarrow (2003) discuss a model of triggering strike-slip faults by formation and rise of K-granitic melts from the base of the lithosphere.

Due to the ascension of magma, we suggest in this review study that the instability localizing the faulting comes from rheological deep-seated levels of the wrench zone. During further development, the instability uprising magma focuses fault motion in the thermally weakened mantle lithosphere or lower crust. Based on the rheological contrasts, there may be further potential mechanisms for the inception of major strike-slip faults, e.g. reactivation of major rift-related normal faults as strike-slip faults (Hsiao et al., 2010), strike-slip faults related to transtension in rifts (e.g., East African rift system; Ring et al., 1992; Acocella, 2014) and melt-induced strain localization and lateral flow of anatetic crust to form extrusion faults (Rosenberg et al., 2007a).

7. Discussion and conclusion

7.1. Recognition of exhumed strike-slip fault zones

Exhumed strike-slip faulting is a deep crustal expression of lithosphere deformation. In particular, crustal-scale exhumed strike-slip faults involve complex processes of origin and structure, as well as thermal and mechanical feedbacks. There is significant divergence between natural examples and models regarding thermal-weakening feedbacks. Observations suggest that the strike-slip faults were already weak (e.g., Rutter et al., 2001) when they initiated as ductile shear zones at deep crustal levels. Oblique-slip faults represent vertical structures that potentially penetrate the weak lower-crust and/or mantle (e.g., Sylvester, 1988). As noted above, faulting and exhumation processes and products change with depth and type of material being deformed. The structure and rheology of strike-slip fault zones vary with depth, such that strong temperature- and rate-dependent ductile behavior in the lower part of the crust and mantle is transitional to pressure-dependent frictional behavior predominate in the brittle upper crust (Alsop et al., 2004; Handy et al., 2007). Although some basic common mechanisms can be identified from multiple models, individual models do show a wide range of structures and potential origins of strike-slip faults. The increase in temperature with depth and localization is probably one of the most important factors influencing crustal viscosity, as has been recognized in several studies of post-seismic deformation (e.g., Riva and Gover, 2009; Yamasaki and Houseman, 2012). The heated lithosphere enhancing localization and viscous weakening is evidenced by thermomechanical models of a multi-component system (Chéry, 2008).

Less attention has been paid to major orogen-scale exhumed strike-slip faults with hot-to-cold contacts, which commonly develop from ductile shear at depth to brittle displacement at shallow crustal levels. These studied locations of large strike-slip faults at the pluon/meta-morphic complex/country rock boundary or following internal contacts suggest that these domains represent favorable sites for subsequent development of strike-slip faulting. Hot-to-cold contacts are better known in ductile low-angle normal faults at the top of MCCs. The principal effect of a hot-to-cold contact in the crust and/or mantle lithosphere is the change to lower yield strength on the hot block along a strike-slip fault. The weak nature of the lower crust and the deep mantle lithosphere also implies: (1) a potential root of a strike-slip fault is a listric
fault at the base of the lithosphere, and (2) potential lithospheric-scale folding of the involved lithosphere close to the major strike-slip fault (Fig. 11D).

7.2. Thermally enhanced rheological weakening of exhumed strike-slip faults

As we highlighted in this review, in natural examples, there is a broad consensus that granitic plutons and metamorphic (core) complexes are often spatially and temporally related to crustal-scale strike-slip fault zones (e.g., Hollister and Crawford, 1986; Genser and Neubauer, 1988; Speer et al., 1994; Neubauer et al., 1995; Vigneresse, 1995a, 1995b; Wang and Neubauer, 1998; Neubauer et al., 1999; Vaughan and Scarrow, 2003; Rosenberg, 2004; Weinberg et al., 2004; Rosenberg and Schneider, 2008; Cao et al., 2011b, 2011c). Pre- and syn-kinematic partial melting and granite intrusions occur along the axis of strike-slip faults (e.g., Rosenberg and Schneider, 2008). Studies in deeply exposed outcrops have shown that strike-slip shear zones have been traced from greenschist grade all the way to the granulite-grade metamorphism, and well into the lower crust, widening with depth (e.g., the San Andreas Fault system; Henstock et al., 1997; Parsons, 1998; Titus et al., 2006). The metamorphic zonation is parallel to the trend of the strike-slip fault zone and the metamorphic grade increases towards the axis of the fault zone (e.g., Tauern metamorphic complex; Genser and Neubauer, 1988; Neubauer et al., 1999; and Alpine fault, Warr and Cox, 2001). The peak conditions of the Oligocene Tauern metamorphic overprint reached amphibolite facies along an axis that has shifted towards the south of the central axis and greenschist facies along the margins (e.g., Frank et al., 1987; Hoinikes et al., 1999; Neubauer et al., 1999).

Plutonic intrusions and hot metamorphic bodies bounded with strike-slip fault zone at depth lead to heating and weakening, which is the driving mechanism for shear zone initiation at rheological boundaries (Fig. 14). Lithospheric heating by mantle upwelling and related magmatic intrusions significantly contributes to localization and weakening of the lithosphere has been addressed in many studies (e.g., Handy et al., 2001; Buck, 2004, 2006; Bialas et al., 2010). Hollister and Crawford (1986) were the earliest to propose that there is a causal relationship between large-scale deformation and intrusion (melting) in orogenic crust. They suggested that an intrusion weakens the lower crust significantly during orogeny, which increases strain rates and augments the exhumation rates of crustal blocks confined between magmatic rocks bearing shear zones.

Hot geological bodies (e.g., granitic intrusions and warm metamorphic complexes) are juxtaposed with cold country rocks on the other side of the strike-slip fault (Fig. 14A, B). This results in shear concentration along such contacts due to thermally enhanced rheological weakening when external kinematic conditions are appropriate. Rheological weakening by heterogeneities associated with magma intrusions and metamorphic complexes may therefore induce a localization of strain within and near the plutons or metamorphic complex and favor the development of a strike-slip fault zone. This mechanism has recently been corroborated by observed structures at the rift system or MCC (e.g., van Wijk et al., 2001; Kendall et al., 2005; Schmeling, 2010). Alternatively, the thermal activity has been regarded as a possible effect of shearing and ascribed to sheared heating (e.g., Rice, 2012).

7.3. Feedback mechanisms within exhumed strike-slip fault zone

Spatial and temporal relationships in oblique plate boundary deformation with transtension and transtension are often related to the interaction between deep lithosphere and large-scale strike-slip faulting (e.g., Molnar, 1988; Teyssier et al., 1995; Spotila et al., 2007) (Fig. 3). These relationships also imply a wide zone of deformation across major strike-slip fault (e.g., Fitch, 1972; Vauchez and Nicolas, 1991; Fossen and Tikoff, 1993; Tikoff and Teyssier, 1994; Teyssier et al., 1995; Dewey et al., 1998). Weakening by magmatic intrusions and hot metamorphic complexes associated with continental strike-slip faults have an important effect on the thermal state and thus the rheological behavior of the lithosphere as already discussed. The other side in a feedback mechanism, was strike-slip kinematics and associated deformation which poses a major control on the pattern of magma emplacement and metamorphic complex exhumation (e.g., Rosenberg, 2004; Schmeling, 2010). Our review results provide a new conceptual framework for interpreting initiation on rheological boundaries of exhumed strike-slip fault zones. The initiation of strike-slip faulting is considered to be resulted from upwelling of the convective asthenosphere, which leads to a modification of the thermal structure of the overlying lithosphere. Specifically, we argue that the initiation of strike-slip faults at depth is primarily controlled by the rheological weakening contrast in the lithosphere and by releasing oversteps in such systems.

7.4. Exhumed strike-slip faults and orocline syntaxes

Continental indentation tectonics is a process that described how rigid blocks indent into weak orogens and result laterally escaping blocks confined by transient strike-slip faults (e.g., Tapponnier and Molnar, 1976; Ratschbacher et al., 1989, 1991). As described above pluton- and MCC-controlled strike-slip faults are often associated with oroclines and syntaxes. These syntaxes and oroclines evolve often at the tip of indenting lithosphere at the edges of plates (e.g., Pamir at the northwestern edge of the Indenting India; Stibner et al., 2013). In these cases, it seems to be the melting of lower crust, which creates all the magma, which is then smeared into the associated strike-slip faults like the Karakoram and Chugach faults. The oroclines and associated large-scale strike-slip faults develop, therefore, during oblique plate convergence and collision and/or post-collisional processes, when oceanic subduction or final oblique continental subduction partitions into orthogonal shortening at the orogenic front and about orogen-parallel strike-slip faults.

Overall, we have reviewed recent advances in the study of the structure and evolution of exhumed continental strike-slip fault zone systems. We have emphasized the importance of the interplay of these plutonic intrusions and hot metamorphic bodies bounded with strike-slip fault zone at depth lead to heating and weakening. Our review results provide a new conceptual framework for interpreting initiation of exhumed strike-slip fault zones on thermally enhanced rheological weakening along hot-to-cold contacts deep within the crust and mantle lithosphere. However, our understanding of the mechanics and processes of exhumed major strike-slip faulting is still limited. Further progress in understanding the properties and evolution of strike-slip faulting will require an integrated approach involving many different disciplines (from field geology combined with detailed microstructural investigation, laboratory experiments, geophysics, hydrogeology, geochemistry, and numerical and experimental modelling).

Acknowledgements

We acknowledge detailed constructive reviews by Walter D. Mooney, Basil Tikoff, Vasileios Chatzaras, one anonymous and encouragement by the editor, Prof Carlo Doglioni. This work is financially supported by the Austrian Science Fund (FWF), grant nos. M-1343 and P28313-N29 to SC and P22110 to FN, National Natural Science Foundation of China, grant 41472188, and MSFPMR201406 of MOST Special Fund from the State Key Laboratory of Geological Processes and Mineral Resources, China University of Geosciences (Wuhan).

References


